Modelling glacial inception of Antarctica during the Eocene-Oligocene boundary

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Abstract

The Antarctic glaciation during the Eocene-Oligocene boundary is one of the quickest changes in the climate system observed in the geological record. CO_2 is assumed to be the major cause of glaciation,[1] although the uncertainty is high. Here we will study the impact of CO_2, the mass balance, bedrock deformation, sea level changes and vegetation on the climate system during the Eocene-Oligocene transition, especially on the East Antarctic ice sheet. We do this with a coupled Zonally averaged Energy Balance Climate Model and a 1-D Ice Sheet Model. The ablation has a large impact on the climate system, therefore we have determined the ablation parameter C_{abl} = -30. Bedrock deformations cause large fluctuations in ice volume, on the other hand, sea level change causes only small fluctuations. The bedrock relaxation time τ_b has also a large impact on the ice volume. We find that vegetation has an impact on the glacial inception of Antarctica. Only a small fraction of the surface has to be covered with forests to delay the inception with hunderds of ppmv CO_2. We estimated that with 10% or less of the surface covered with forests, it still results in a plausible CO_2 concentration of glacial inception.[1,2] With more than 10% forests, the ice sheet develops only with very low CO_2 concentrations, or it does not develop at all. The temperature of the warmest month is high enough to sustain forests at the margins of the continent. With calculations of the model, we conclude that the mass balance and bedrock deformation are important on the glacial inception. Vegetation however, has the largest impact on the glacial inception of Antarctica.

Contents

1 Introduction

Mass changes of ice are influenced by climate change. They are dependent on many different variables, such as the albedo, precipitation, insolation, CO_2, the atmospheric lapse rate, the ocean circulation and many others. Here we focus on the glacial inception of Antarctica. This inception took place during the Eocene-Oligocene transition.

The Eocene is an epoch that lasted roughly from 56 to 34 million years ago. It is the second epoch of the Paleogene period, followed by the Oligocene epoch, which lasted from 34 to 23 million years ago. The Eocene was characterised by warm temperatures, high CO_2 concentrations and high sea level. In addition, the land configuration was different than today. Antarctica and Australia were still connected, just as South America and Antarctica. The Tethys sea way was closed and North America was not yet connected to South America. Large parts of West Antarctica were possibly below sea level, due to the high sea level at that time. Ocean circulation was therefore very different than it is nowadays. This had a major impact on the heat distribution caused by the ocean. Especially Antarctica received more energy by the oceans than today.[3] In the present climate, Antarctica is thermally isolated, because of ocean currents going around Antarctica. The topography of Antarctica was also different. The present topography is completely dominated by a kilometers thick ice sheet, which has been there for more than 30 million years, and surpresses the underlaying bedrock. During the Eocene, ice free conditions prevailed and mountain ranges dominated the landscape. These mountain ranges are the places of first glaciaton.

From the Oligocene and after, the world had become an icehouse, due to ice present on Earth. The Eocene-Oligocene boundary was therefore an important transition. CO_2 concentrations during the Eocene were much higher than today. The CO_2 concentrations during most of this epoch were between 750 and 900 ppmv. The uncertainty around the Eocene-Oligocene boundary is large (see figure 1). Figure 1 shows CO_2 proxy data,[2] which includes multiple ways to reconstruct the past CO_2, most of them are marine and terrestrial proxies. The figure shows that acceptable CO_2 concentrations are within the range of roughly 500 and 1250 ppmv.

The uncertainties in CO_2 concentrations during the Paleogene in previous work were large.[2] Previously the concensus was that the tectonic movement of Antarctica caused the glaciation during the Eocene-Oligocene transition,[4] but a more recent study suggested that the major cause of the glacial inception of Antarctica was in fact the drop in CO_2 concentration.[1] The thermal isolation of Antarctica played a major role in the first ice development, but it was not the most important process. Tectonic movements only caused a delay in the glacial inception. Figure 2 shows the ice volume of the Antarctic ice sheet in the case that that the Drake passage was open and closed, with decreasing CO_2 concentrations on the right axis. Figure 2 shows that the glacial inception happened at lower CO_2 concentrations if the Drake Passage were still closed.[1] The glacial inception would still happen, but it would be with lower CO_2 concentra-

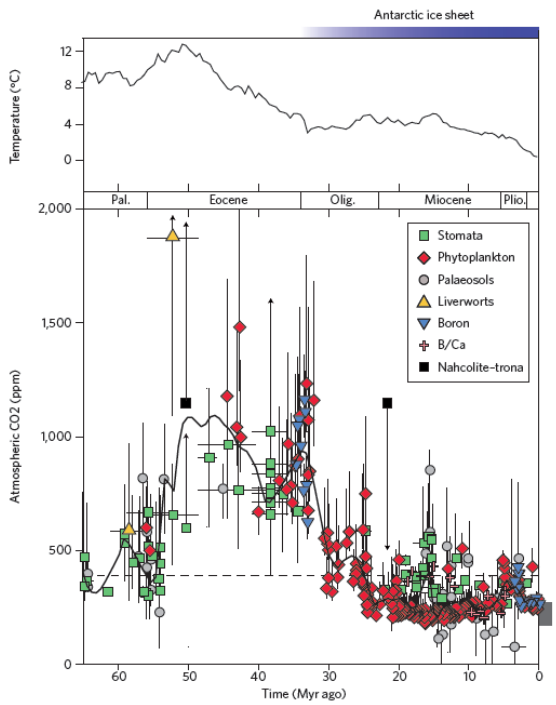

Figure 1: *CO_2 concentration approximations for the Cenozoic era.*[2]

tions. Around 2.4 times the preindustrial CO_2 concentration, roughly 650 ppmv CO_2, the total amount of ice volume was almost the same as when the Drake Passage was open. Determining the CO_2 concentration around the Eocene-Oligocene boundary is part of this research.

The Eocene-Oligocene boundary was characterised by the sudden development of ice on Antarctica. Around the boundary, the CO_2 concentration passed a certain threshold. This leads to mechanisms that enhanced ice growth and caused lower temperatures, such as the ice-albedo feedback and the ice sheet height feedback. When the Milankovitch parameters (precession, obliquity and eccentricity) were optimal for ice development, it caused a very fast ice growth on mainly East Antarctica. Initially there were many small ice sheets on the continent, most of them did not reach the sea.[1] The volume of the ice sheets were dependent of the Milankovitch cycles, with ice growth during periods with low insolation at high latitudes, caused by obliquity. Around a CO_2 concentration of 780 ppmv, the ice sheets became large enough to merge with each other to form the East Antarctic Ice Sheet.[1] This concentration has a large uncertainty.[5]

The climate of the past can be determined in different ways, such as tree rings, ice cores and benthic sea records. Ice cores and sea sediment

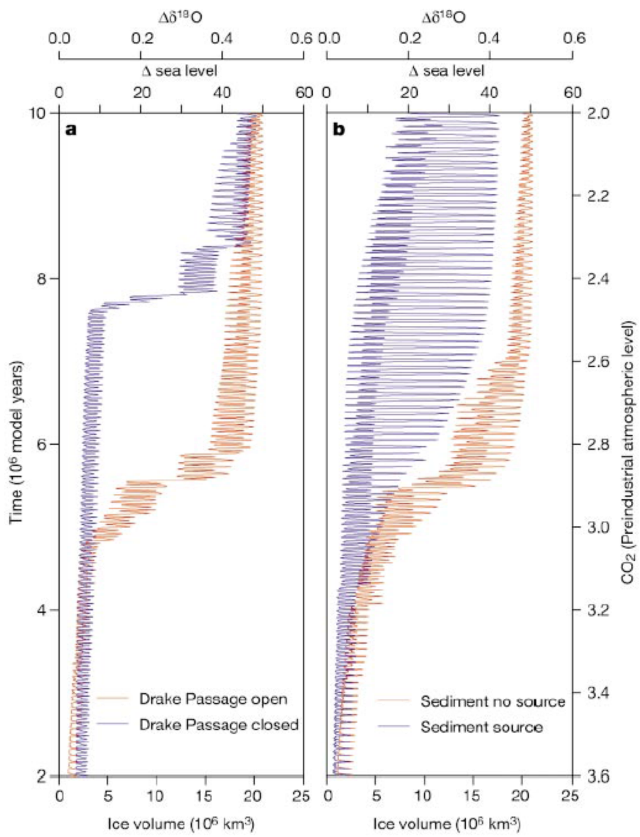

Figure 2: *Ice volume difference between an open and closed Drake Passage.*[1] *On the right axis the CO_2 concentration and on the left axis the modelled time.*

records are the most common for paleoclimate research. Ice cores do not go back in time long enough to be relevant for the glacial inception of Antarctica, so they will not be discussed here. Marine sediment records go back a sufficient amount of time to be useful.

Benthic sea records are records from the bottom of the ocean, from the benthic zone. The sediment layers are stratified, with the younger layers on top of older layers. If the place for a record is well chosen, the deepest sediment layers are tens of millions of years old, old enough to provide information from the Eocene-Oligocene boundary. The sediment layers contain calcite shells from foraminifera. They contain information about the temperature of the ocean and ice volume of that time. The chemical composition of the shells are dependent on the environment they have lived in, and can therefore be used for determining the circumstances of that place at a certain time. When a measurement is done, the isotope ratio of different elements can be determined for many time periods. The most useful ratio is the δ^{18}O. δ^{18}O is defined as (see equation 1).

$$\delta^{18}O = \left[\frac{(\frac{^{18}O}{^{16}O})_{sample}}{(\frac{^{18}O}{^{16}O})_{standard}} - 1 \right] \cdot 1000‰ \tag{1}$$

The δ^{18}O is a fraction between the measured ^{18}O and ^{16}O, compared to the standard ratio. This fraction is a measurement for the temperature and ice volume during that time. In the oceans δ^{18}O can vary between -2‰ (warmest/least ice) and 4‰ (coldest/most ice). Water with a ^{16}O oxygen isotope has a smaller mass than a water molecule with a ^{18}O atom. So $H_2{}^{16}O$ vaporises slightly easier than $H_2{}^{18}O$. Some of the vaporised $H_2{}^{16}O$ precipitates on an ice sheet and will not return to the ocean on a short term, taking ^{16}O out of the ocean. Diffusive processes are temperature dependent, so when the shells are produced, the δ^{18}O in those shells is a characteristic of the sea temperarature. Therefore, $\frac{^{18}O}{^{16}O}$ present in the sea sediment records will be lower when it is warm and when only small volumes of ice are present on Earth, and higher when it is cold and high volumes of ice. As a result, it is a measure of the temperature and ice volume.

This research will focus on better understanding the influence of the mass balance, vegetation, isostasy, sea level and CO_2 concentration on the Antarctic glaciation. Due to the high uncertainty of CO_2 in previous work,[5] it is important to better estimate the concentration for the Eocene-Oligocene boundary. Tuning parameters are often used to test the sensitivity of the ablation and bedrock. We will test these tuning parameters in order to find the values that produce the most comparable results.[1,2] Vegetation is suggested to be of large influence on the climate system.[10] We will therefore do tests with different vegetation present on Antarctica. Vegetation will have its effect on the albedo, which will often result in an increase in temperature. Ice volume changes and mass balance components will be used to test the ice sheets, but also other quantities, like temperature. We will also make cross sections of the East Antarctic ice sheet. We use a coupled Zonally averaged Energy Balance Climate Model (ZEBCM)[6] with a one dimensional Ice Sheet Model (ISM).[7] The coupled model has already been used for modelling the glaciation cycles of the last 800 kyrs.[8] The sensitivity of the climate system to variations in the CO_2 concentration has already been tested for this time period, just as the ocean overturning and the climate-ice sheet feedbacks.[8] Most of the physical processes that play a role in the glacial inception are part of the model. This model is therefore useful for sensitivity experiments on Antarctica. We conclude which of the processes are relevant during the Eocene-Oligocene transition, after we determine the sensitivity of the mass balance, bedrock, vegetation and CO_2 to changes in the climate system.

We will discuss the ZEBCM and ISM in more detail. We will also discuss how both models work, and how the relevant physical laws are parameterized. Then we will discuss the theoretical background about some of the most important physical processes. These processes have an important role in the mass balance of the ice sheets. The processes described are calving, isostasy and vegetation. Then the results will be shown. First we will test the model, then the equilibrium time experiments, mass balance experiments, bedrock and bedrock relaxation time tests and vegetation. Finally, we will discuss the results and make a conclusion.

2 Model

This research focusses on the use of models. We will use the models to test the sensitivity of the ice sheets, mostly the East Antarctic Ice Sheet. The two models used are the Ice Sheet Model and the Zonally Averaged Energy Balance Climate Model. These models have been coupled.[8]

2.1 Zonally Averaged Energy Balance Climate Model

The first part of the coupled model, is the Zonally Averaged Energy Balance Climate Model (ZEBCM). This is a model that divides the Earth in zonally equal bands of 5° in latitude. So in total there are 36 bands across the Earth. Within these bands, the climate is averaged. The model has many physical processes incorporated. First of all, there is CO_2, obviously a very important quantity. Some of the others are: turbulent heat transport, meridional heat transport, ocean circulation, insolation, local topography, seasonal cycle and vegetation. The vegetation has its influence on the albedo. Because it is a zonally averaged model, every latitudinal band has a ratio between the different types of vegetation. The same is true for snow cover and land-sea ratio, but both will not be explicitly varied here. Our model assumes that Antarctica is completely covered by grass (which will be changed as an experiment), but at places with glaciation, the vegetation is gone and the snow albedo prevails. The albedo $\alpha = 0.15$ for grass and forests. When snow falls on grass, the albedo quickly changes to that of snow, but when snow falls on a forest, the canopy reduces the albedo. We use the albedo[6] for grass covered with 'cold' snow $\alpha_1 = 0.80$ and for 'warm' snow $\alpha_2 = 0.40$, in which 'cold' refers to $T < 263$ K and 'warm' refers to $T > 273$ K. For forests $\alpha_1 = 0.40$ for 'cold' snow and $\alpha_2 = 0.30$ for 'warm' snow. For 263 K $\leq T \leq$ 273 K, the albedo is given by equation 2. Here we assume that the albedo declines linearly.

$$\alpha_{grass} = \alpha_1 + [\alpha_2 - \alpha_1]\frac{T - 263}{10} \tag{2}$$

The model can use CO_2 in two different ways. First of all we can use the CO_2 to force the model. This is useful when testing at certain constant CO_2 concentrations. Secondly, we can use the model in an inverse way. The model will then run with δ^{18}O from sea records from previous work[9] and other input records such as insolation, but this method will not be used. In this research we will force the model with either CO_2 measured from ice cores, or by a prescribed CO_2 concentration.

Figure 3 shows a schematic overview of a zonal band in the ZEBCM. At the top of the atmosphere, the incoming short and longwave radiation

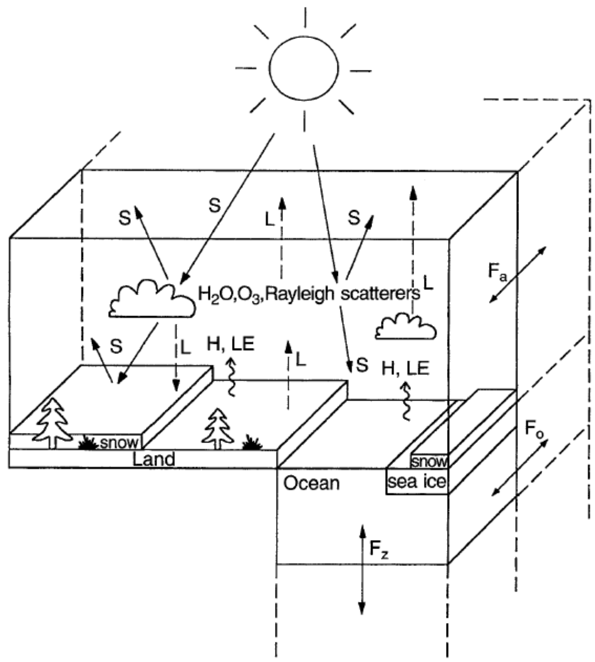

Figure 3: *Schematic overview of a zonal band in the ZEBCM.*[6] *S is shortwave radiative flux. L is longwave radiative flux. H is the turbulent heat flux. LE is the latent heat flux at the surface. F_a is the total heat flux in the atmosphere. F_o is the total heat flux in the ocean, and F_z is the vertical heat flux between upper ocean layer and deep ocean. The ZEBCM also contains the snow/land ratio, ocean/sea ice ratio and the vegetation ratio.*

are evaluated. At the lowest atmosperic layer, the radiative and turbulent fluxes are exchanged with the surface. The fluxes are expressed in terms of the zonally averaged temperature T_a. Most of the shown fluxes will not be discussed any further. The snow fraction, sea ice fraction and vegetation are all included.

2.2 Ice Sheet Model

In this research we use a one dimensional Ice Sheet Model (ISM)[7] to test how the ice sheets would react to changes. It consists of five ice sheets, namely: the East Antarctic Ice Sheet (EAIS), West Antarctic Ice Sheet (WAIS), Eurasian ice Sheet (EAS), North American ice Sheet (NAM) and the Greenland ice Sheet (GRL). The ISM assumes that the continents are rotational symmetric and have in fact a cone shape. This way we can simplify the ice sheet to one dimension. The ice sheets are thick in the centre of the continent, and thin at the margins. We can question to what extent the continents are cone shaped, but especially the EAIS is well approximated by a cone. The EAIS is also the largest, and most important ice sheet of this research. Ice shelves will not form in this model. Therefore all ice that reaches the ocean will calve instantaneously.

The ISM calculates the ice sheets on a grid. On each grid point the model calculates all the relevant variables, such as bedrock height, ice height, temperature, accumulation and ablation. The EAIS is divided into 120 gridpoints, with a gridsize of 20 kilometers. The height of the background before glaciation is 1410 meters above Present Day (PD) sea level, which is in the centre of the continent. After each gridpoint the height of the bedrock decreases by 20 meters. Thus East Antarctic bedrock has a constant decreasing slope, all the way down to the last (120th) gridpoint, which is at a height of 950 meters below PD sea level. After glaciation, isostatic effects will deform the bedrock. The weight of the ice pushes the bedrock down, causing it to sink, and potentially become underneath the sea level. Increasing heights in front of the ice sheets and gravitional effects are not taken into account in the model, only the local deformation caused by the weight of the ice sheet. This process is important for a potential connection of the ice sheet with the ocean, which can accelerate the mass loss.

An important part of the ISM, is the mass balance. The model calculates the mass balance with precipitation, ablation and calving. Precipitation depents on the temperature, as well as the total amount of ice present on the continent. The temperature dependency is a result of the Clausius-Clapeyron equation, which states that the precipitation increases 4% per K. If the ice sheet expands, it will receive relatively less precipitation in the centre, so it is a function of the radius of the ice sheet. Combining this with a reference and a tuning parameter for the model, will result in a precipitation (P) equation.[8]

$$P = P_0 \cdot e^{0.04T - R/R_C} \tag{3}$$

P_0 is the present day reference value, and R_C is a tuning parameter, which can be varied for a sensitivity test, or can be kept constant to match measurements. T is the temperature and R the radius of the ice sheet.

Ablation (M) and calving (C) are the physical processes that cause the ice sheet to lose mass. Calving is the process of mass loss due to ice berg formation when the ice sheet reaches the ocean. The ablation is dependent on the temperature (T), albedo (α), incoming radiation (Q) and some tuning parameters. The following ablation equation is the result if we combine these parameters.[8]

$$M = \frac{1}{100}[10T + 0.513(1-\alpha)Q + C_{abl}] \tag{4}$$

In equation 4, C_{abl} is a tuning parameter, 10 and 0.513 are reference parameters. The 10 and 0.513 arise from fitting with the present. Q is the average insolation for the particular ice sheet. The parameter C_{abl} de-

termines wheter or not there is ablation, it is therefore useful to do tests with C_{abl}.
The net mass balance (B), is given in equation 5, with calving (C) included.

$$B = P - M - C \tag{5}$$

The flow of ice can now be described by the continuity equation, given by equation 6.

$$\frac{\partial H}{\partial t} = \frac{1}{r}\frac{\partial}{\partial r}(rH\bar{U}) + B \tag{6}$$

H is the thickness of the ice, r the distance from the centre, $\bar{U}$ the mean horizontal velocity, and B the mass balance as described in equation 5.

In this study we test the sensitivity of the mass balance of Antarctica, by means of varying the parameter C_{abl}.

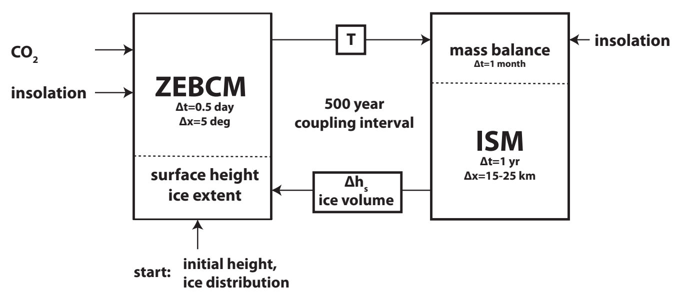

Figure 4: *Overview of the forced coupled model.*[8]

Figure 4 shows a schematic overview of the coupled model. We will force the ZEBCM with CO_2 and insolation. After 500 years, the ISM receives temperature data from the ZEBCM. The ISM calculates and the ZEBCM receives the new data about the ice sheets from the ISM, as well as new CO_2 and insolation data. This data will then be used to calculate new temperatures, with which the ISM calculates the ice volume. This will be repeated through time.

3 Mass balance of the early Antarctic ice sheet

3.1 Calving

When the ice sheets are large enough to reach the ocean, they start to lose mass not only by melting, but also by calving. Calving happens when the ice sheet reaches the ocean and slowly starts to break down. Therefore calving accelerates the mass loss. When a warm period takes place, the ice sheet will retreat. The ice sheet will often lose its connection to the ocean, and the sea level will rise due to the decreasing ice volumes. The initial response of the ice sheet, just after losing connection to the ocean, is to grow again, because there is no more mass lost due to calving. It is therefore possible to see a short peak in ice volume, even when the temperature and CO_2 concentrations are rising. Of course, if the temperature continues to rise, this peak will fade away.

3.2 Isostasy

The same effect as described in section 3.1 can take place due to isostasy. An ice sheet present on the continent will push the bedrock down. This causes the ice sheet to be less high in the atmosphere and it will gain contact with the ocean more rapidly. When the ice sheet loses mass, the bedrock will slowly come up again. This process is called isostasy. It causes the ice sheet to grow again, due to being higher in the atmosphere. When the ice melts, in combination with isostasy, it can lose contact with the ocean. When this happens, there is suddenly less ablation, causing the ice sheet to grow until it reaches the ocean again. When the melting keeps increasing, the described process will be completely compensated. The lithosphere and asthenosphere will deform when ice is present. Equation 7 describes the process.[8]

$$\frac{\partial b}{\partial t} = -\frac{1}{\tau_b}\left(\frac{H}{k} + b - b_0\right) \tag{7}$$

τ_b is the relaxation time of the asthenoshere. The relaxation time is the timescale of the bedrock to deform. H is the ice thickness, b and b_0 the height and initial height of the bedrock and k is the density ratio between the ice and the bedrock, which is kept constant at $k = 3$. The initial height of the bedrock is given by $b_0 = H_{bc} - (\frac{db}{dx})x$, with $H_{bc} = 1410$ m above present day sea level and x the distance along the flowline. When the relaxation time τ_b is small, the bedrock reacts fast to ice volume changes. When τ_b is very small, the bedrock reacts almost instantaneously. Large fluctuations in ice volume can be caused by a large τ_b. This is due to the long equilibrium time of the ice sheet when τ_b is big, because the bedrock deforms slowly. Gravitational and elastic effects will be neglected.

3.3 Vegetation

Before the glacial inception of Antarctica, there was vegetation present at the continent. This vegetation could have a significant impact on the timing of glaciation. This was due to the lower albedo α of the different kinds of vegetation when it was covered with snow. Grass was covered completely for small amounts of snow, while the canopy of forests reduce the albedo and slowing down the cooling (see section 2.1). A forest can therefore slow the time of first glaciation significantly.[10] In previous work, the vegetation has been modelled with an atmospheric model combined with a vegetation model to give an idea how this might have changed with different CO_2 concentrations[10] (see figure 5).

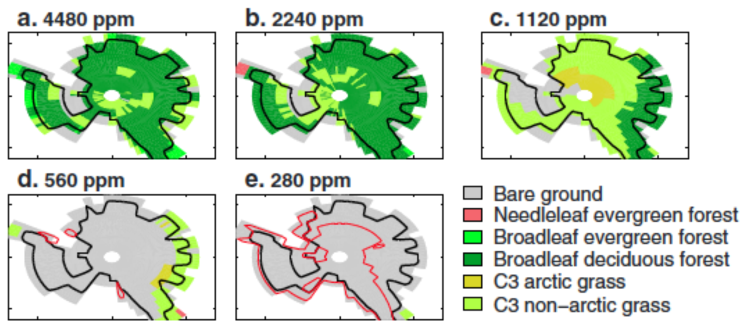

Figure 5: *Modelled vegetation for Antarctica for different CO_2 concentrations.*[10]

Figure 5 shows the vegetation modelled in previous work for different CO_2 concentrations. The red contours show the everlasting snow cover. In figure 5 it is shown that around 1120 ppmv CO_2 there were still forests present at the margins of East Antarctica. They came to the conclusion that the CO_2 concentration to start glaciation for Antarctica dominated by forests, would be much lower than previously assumed. He determined that the concentration would be lower than 280 ppmv, while for bare ground and tundra it would be between 560-1120 ppmv. Therefore, forests were an important factor that slowed ice development down. They concluded that the transition from forests to tundra have been most likely before the Eocene-Oligocene boundary,[10] which is in accordance with previous work.[1] The transition from forests to tundra may have been due to too low summer temperatures. Forests can withstand low temperatures, but when the summer temperature is too low and therefore the growing season is too short, forests cannot survive. The lowest mean temperature of the warmest month[11] with which forests can still grow, is approximaty 8 °C. Below this temperature, tundra vegetation takes over. In this study, we are going to test the impact of vegetation on the glacial inception of Antarctica, especially the difference between forests and grass on the glacial inception.

4 Results

In this research, the emphasis will be on sensitivity experiments of the East Antarctic ice sheet. We will study the mass balance, isostasy, vegetation and CO_2.

4.1 Modelling of ice and temperature during the last 410 kyrs

First, we do a test run. We test if the glaciation cycles of the last 410 kyrs could be modelled correctly. All the large ice sheets of the recent past are taken into account in the model. The model is forced with CO_2 data (see section 2.1). The purpose of this experiment is to test if all the different major ice sheets work properly. We let the model run a spin-up for 90 kyrs with CO_2 concentrations of 240 ppmv. The CO_2 concentration was kept constant 10 kyrs from the present, in order to model the Holocene average in a correct way. The result of this experiment can be seen in figures 6 and 7.

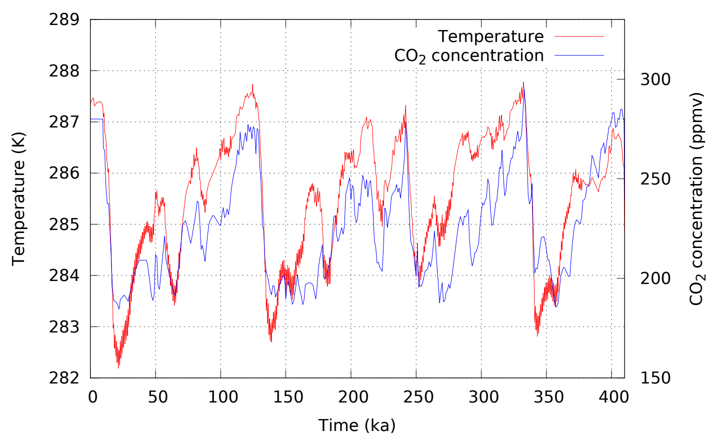

Figure 6: *Mean temperature of the Earth of the last 410 kyrs. The 100 kyrs glacial cycles are clearly visible. The used CO_2 concentration is also shown.*

Figure 6 shows the mean temperature on Earth, as well as the CO_2 concentration. The glacial minima and maxima are clearly visible, which are in agreement with observations.[12]

Figure 7 shows the ice volume of the different major ice sheets. The ice sheets react quickly to changes in CO_2 concentrations. The ice volumes as illustrated in figure 7 grow and decay almost as expected.[12] The NAM grows and decays rapidly. The EAS does the same, but in a lesser extent. Both of them can dissapear in glacial minima. The model calculates that the ice sheets would almost disappear (less than 3 meter SLE ice volume) in the Holocene. The GIS and WAIS have the volume which is in agreement with observations,[12] just as the EAIS. The EAIS loses some mass after a glacial maximum, but after approximately 20 kyrs it is back in equilibrium. This may be caused by the ice sheet gaining or losing contact with the ocean (see section 3.1 and 3.2).

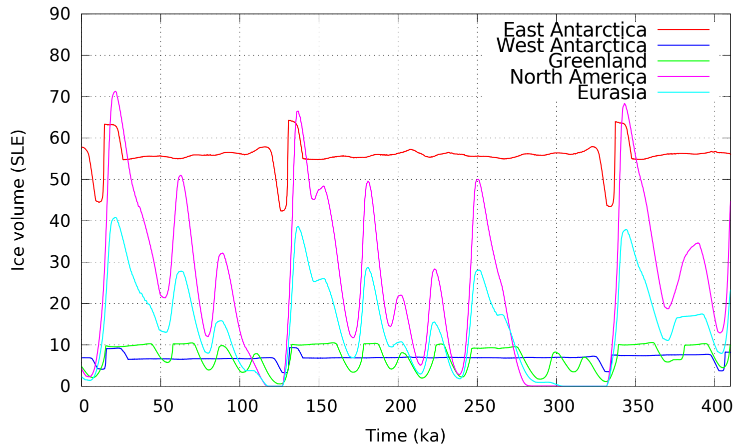

Figure 7: *Ice volume of the five major ice sheets of the last 410 kyrs. The glaciation cycles are clearly visible.*

4.2 Determining the equilibrium time

An important experiment, with which the results will be used in most of the other experiments, is the testing of the equilibrium time of the Antarctic ice sheet (see figure 8). Figure 8 shows how long it takes for the EAIS to grow to its full extent. The CO_2 concentration was kept constant at 600 ppmv, so we now use a forced method. In order to determine the equilibrium time, we have to run the model until the ice volume is constant. When the ice volume does not change in time anymore, the ice sheet is in equilibrium.

The EAIS reaches its maximum just after 50 kyrs and is in equilibrium after about 60 kyrs. As a result, 60 kyrs is the equilibrium time used for the EAIS. In most of the experiments, the CO_2 concentration is changed in a stepwise way. The width of each step is 60 kyrs. Therefore we can assume that the EAIS reaches equilibrium in this time window.

4.3 Reference run

The reference run which we will use to compare results with, is shown in figure 9. The East Antarctic ice volume is shown, together with a stepwise CO_2 concentration, varying between 500 and 1000 ppmv CO_2. The stepwidth of the CO_2 concentration is the equilibrium time as determined in section 4.2. We use C_{abl} = -30, PD insolation, varying bedrock and sea level, τ_b = 3000 years and all of East Antarctica covered with grass when no ice is present. Glacial inception takes place in this reference run

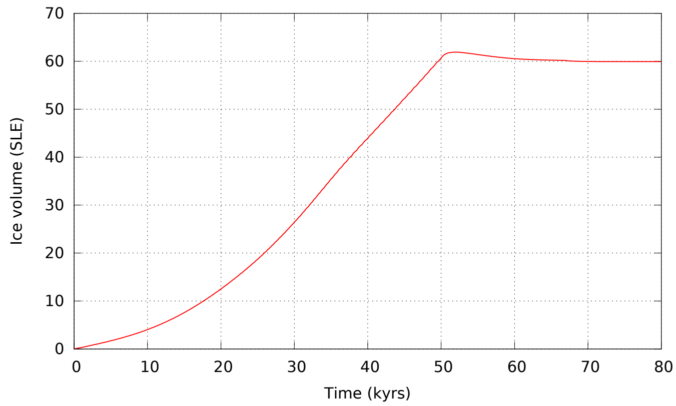

Figure 8: *Equilibrium time test for the EAIS. The ice volume of the EAIS is shown here. This test was done for a CO_2 concentration of 600 ppmv.*

around 650 ppmv CO_2, and the ice sheet starts to dissappear around 850 ppmv CO_2, so a hysteresis is visible.

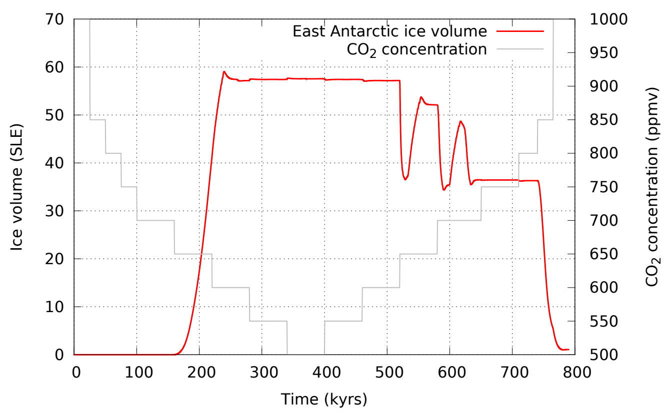

Figure 9: *Reference run for the EAIS. The ice volume of the EAIS is shown here. This test was done for a stepwise CO_2 concentration, C_{abl} = -30, PD insolation, varying bedrock and sea level, τ_b = 3000 years and all of East Antarctica covered with grass when no ice is present.*

4.4 Mass balance experiments

An important part of this report, is testing the mass balance of East Antarctica with circumstances similar to those of 34 million years ago, around the time of glaciatic inception of Antarctica. Hysteresis will be tested as well. To determine the hysteresis, the CO_2 is dropped from 1000

ppmv to 500 with a steplike function and then increased in the same way. A hysteresis is visible if the ice volume increases differently than it decreases. Most of the steps had a stepwidth of the size of the equilibrium time determined in section 4.2. The insolation is kept constant at PD value for °65 NH (Northern Hemisphere) in the ice model, and constant at PD values for all latitudinal bands in the climate model.
The ablation parameter C_{abl} (of equation 4) is changed for this experiment. The ablation parameter C_{abl} is varied between -50 and -10. This parameter is added to the other terms (equation 4), so larger negative values of C_{abl} correspond to less ablation. When there is less ablation, the mass balance becomes more positive, thus the ice sheet tends to grow. Figure 10 shows the ice volume of the EAIS for different values of C_{abl}, combined with the CO_2 concentration, with the ice volumes on the left axis and the CO_2 concentration on the right axis. In figure 10 we can see that at certain CO_2 concentrations the ice sheet grows very rapidly. This is because of feedback loops. When the ice grows, there is a larger part of the continent covered by ice, so the albedo becomes higher. The temperature will drop consequently to cause even more ice growth. In all cases (except for C_{abl} = -10, where nothing happens, which is highly unlikely for the EAIS to not have ice with 500 ppmv CO_2[1,2]), the ice sheet grows to a maximum, declines for a few meters of sea level equivalent and reaches a new equilibrium. These small overshoots are an artifact of the resolution of the model. The model calculates in discrete time steps, so the time step just before the overshoot is still favorable for ice growth, but the next step is not. As a consequence, the model overshoots.

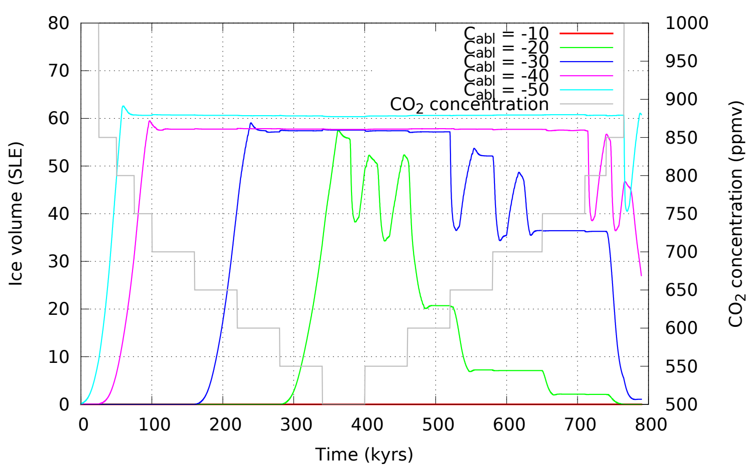

Figure 10: *EAIS mass balance sensitivity experiment, combined with varying ablation parameter C_{abl}, PD insolation for °65 NH and stepwise varying CO_2 concentration. On the left axis the ice volume in sea level equivalents, which corresponds to all colored lines. On the right axis the CO_2 concentration in ppmv, corresponding to the grey background line.*

If we consider the CO_2 concentration of the glacial inception of Antarctica at different ablation parameters C_{abl}, we observe that the ice sheet starts to grow for C_{abl} = -20 at 550 ppmv CO_2, for C_{abl} = -30 approximately at 650 ppmv CO_2 and for C_{abl} = -40 at 850 ppmv. If we compare the results of this test with the estimated CO_2 concentrations of previous work,[1,2] which was estimated around 700 ppmv, we conclude that the value of C_{abl} which produces the most comparable result, is between C_{abl} = -30 and C_{abl} = -40. C_{abl} = -30 has already been used as the default value, and we conclude that it is consistent.

Most of the modelled ice sheets come sooner or later in equilibrium, C_{abl} = -20 and C_{abl} = -30 even have multiple equilibria. A hysteresis is clearly visible. The ice decays with higher CO_2 concentration than needed for initialization. It also has multiple equilibria when the ice decays, which is not the case for ice growth. The most notable feature of the graph, are the multiple peaks when the ice sheet starts to decay. It is visible for all the values of C_{abl} except for C_{abl} = -10. We expect calving to be responsible, which can become zero when the ice sheet loses contact with the ocean, due to the sea level change and/or the bedrock (see section 3.1 and 3.2). We will come back to this in other experiments.

4.5 Bedrock and sea level tests

To test what causes the multiple peaks in figure 10, we are going to keep the bedrock and/or sea level constant, with PD insolation and apply the same stepwise CO_2 concentration function. This experiment was done in order to see what happens to the ice volume when the ice sheet is in contact with the ocean and when it is not. When an ice sheet that is connected to the ocean decays and retreats on land, the mass balance (equation 5) can change sign and the ice sheet can regrow again. This can lead to a reconnection to the ocean, causing increased calving and the ice sheet will retreat again. The bedrock can cause the same phenomenom (see section 3.2). We have tested this for different ablation parameters C_{abl}. Figure 11 shows the calculated ice volumes for C_{abl} = -30.

Figure 11 shows the EAIS ice volume for when bedrock and sea level are constant/varying, combined with the CO_2 concentration. When the bedrock is set to vary, there are still peaks visible after first decay of the EAIS. This is true for both the cases with varying and constant sea level. The sea level only causes the peaks to be delayed, but the first growth and the last decline are still the same. From this we conclude that the sea level does not cause the ice volume fluctuations after first decay. Secondly, we test the case that bedrock is constant and sea level is constant/varying. Both cases are very similar. This is because the ice sheet grows much more rapidly to the ocean than the sea level would fall due to the increased amount of ice on land. The sea level falls only in the order of tens of meters and the ice reaches depths of hundreds of meters (see figure 12). Figure 12 shows the cross section of the EAIS for different times. The sea level and bedrock are set constant in the figure. When the bedrock is constant and the sea level is set constant or set to vary, the described

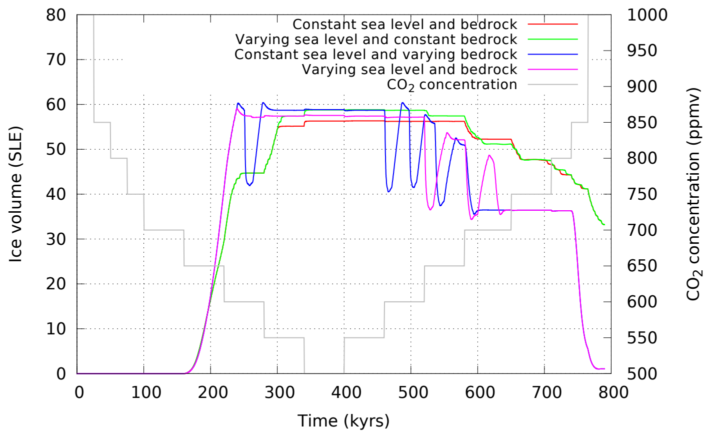

Figure 11: *EAIS ice volume when bedrock and sea level are constant/varying, with PD insolation, C_{abl} = -30 and stepwise function for CO_2. On the left axis the ice volume, on the right axis the CO_2 concentration.*

fluctuations are not visible anymore. From figure 11 we conclude that the peaks after first decay are caused by bedrock fluctuations and not by sea level variations, which is a consequence of isostasy (see section 3.2). This is to be expected, because the bedrock deforms for hundreds of meters due to isostasy, while the sea level varies only tens of meters.

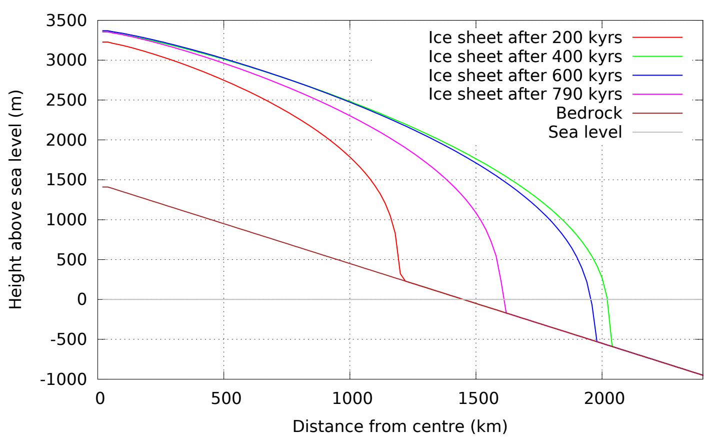

Figure 12: *Cross section of the East Antarctic ice sheet. Bedrock is kept the same as before glaciation. Sea level and insolation have PD values. C_{abl} = -30. CO_2 is varied as before. Cross sections at multiple times have been drawn, just as the bedrock and sea level.*

To give more insight in how far the ice sheet is away from the ocean, we have made cross sections of the EAIS for different cases. One of them is shown in figure 12. We can see that the ice sheet gains contact with the ocean rather quickly. The mean East Antarctic temperature and δ^{18}O of the experiment with the constant bedrock and sea level are shown in figure 13. The temperature and δ^{18}O behave in accordance with the ice development and CO_2 concentration, with a large peak downwards when the ice first starts to form. The mean temperature is low, for all times below the freezing point. As a result, the ice sheet quickly starts to form.

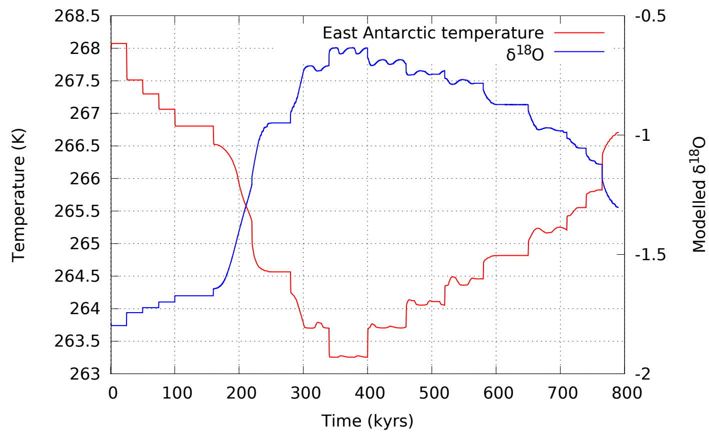

Figure 13: *Yearly mean East Antarctic temperature on the left axis and modelled $\delta^{18}O$ on the right axis for the case with bedrock constant and sea level constant. C_{abl} = -30, PD insolation and stepwise CO_2 concentration.*

4.6 Bedrock relaxation time

To test the sensitivity of the EAIS of the bedrock relaxation time τ_b, we do experiments with multiple values for τ_b. We test τ_b = 1, 3, 5 and 10 kyrs. 3 kyrs is the bedrock relaxation time used in the reference run. Figure 14 shows the ice volume of the EAIS for different values of the bedrock relaxation time τ_b on the left axis and the CO_2 concentration on the right axis.

Figure 14 shows that the variations in ice volume increase with increasing relaxation time τ_b. On the other hand, with small τ_b, the fluctuations are almost gone. This reinforces the idea of section 4.5, that the bedrock is the phenomenom that causes the fluctuations just after first ice volume decay (the peaks visible in figure 10). The variations of the ice volume with $\tau_b = 10000$ years and $\tau_b = 5000$ years are too large to be physically correct.[1,2] We estimate the bedrock relaxation time τ_b around $\tau_b = 3000$ years. In order to get an idea about how the EAIS evolves through time, we show the cross section of the EAIS for the case of $\tau_b = 10000$ years (see figure 15). The figure shows that when the changes in ice volume are sudden, the bedrock did not have time to attain a new equilibrium (e.g.

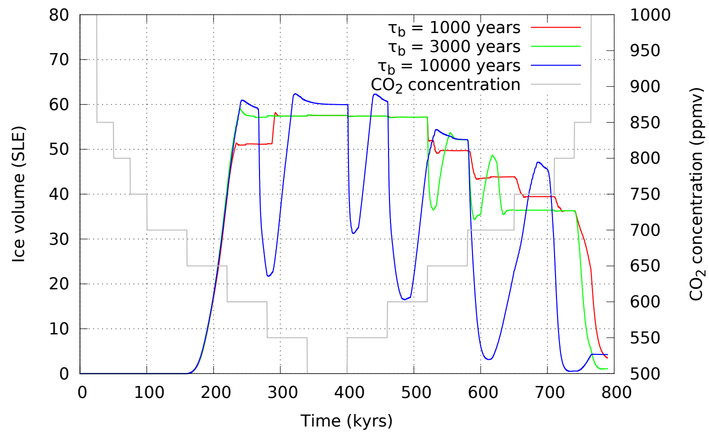

Figure 14: *The bedrock relaxation time for East Antarcitca for multiple values of τ_b. We used the same stepwise CO_2 concentration as before, PD insolation at 65° NH, C_{abl} = -30 and a sea level that can vary.*

the case after 600 kyrs in figure 15). This can have its feedback on the ice sheet. When ice suddenly dissapears, the bedrock rebounds, causing the ice sheet to be less high in the atmosphere than it would be when changes are instantaneously, what its consequence has on the growth of the ice sheet.

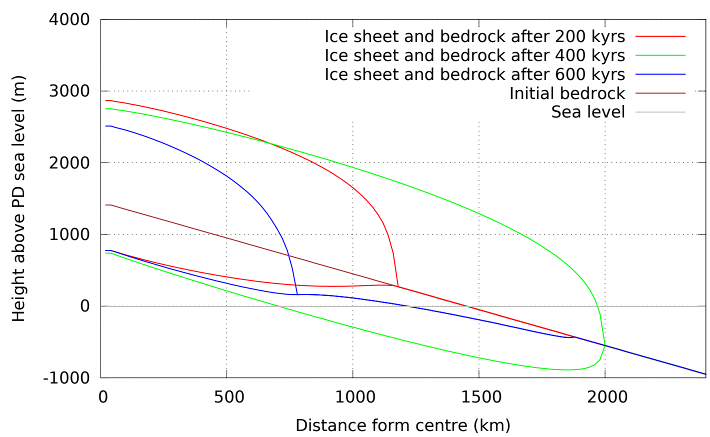

Figure 15: *The EAIS cross section for τ_b = 10000 years. The bedrock deforms much through time, but is not instantaneously deformed to its new equilibrium position.*

4.7 Vegetation

It has been shown that the impact of vegetation on the mass balance can be large.[10] Forests have a lower albedo than snow that falls of grass, as described in section 2.1 and 3.3. Here we test the influence of vegetation on the mass balance. In order to do so, we need to determine the mean temperature of the warmest month, because below a threshold of 8 °C forests cannot survive.[11]

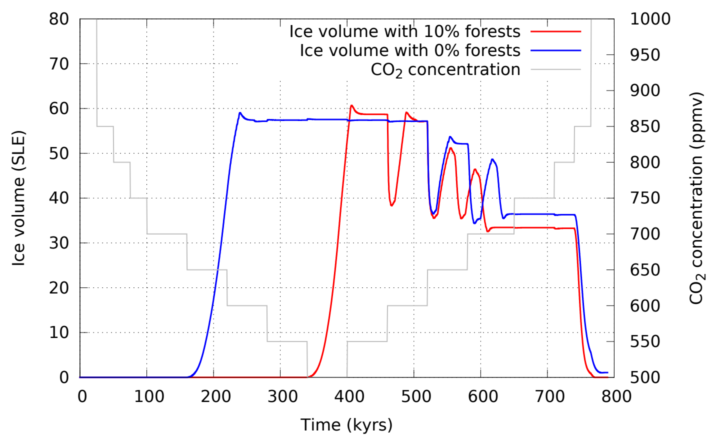

Figure 16: *East Antarcitc ice volume on the left axis and stepwise CO_2 concentration on the right axis. The ratio of forests and grass in places with no ice is changed in the ZEBCM. The graph shows the percentage of the surface covered with forests. Ice growth does not happen with higher percentages, only with 10% and 0% of the surface covered with forests.*

Figure 16 shows that the ice volume of the EAIS is strongly dependent on the vegetation present in Antarctica. We have put the same ratio of forests compared to grass in the model for all the Antarctic latitudinal bands (from 65° to 90°). Only for the cases with the surface covered with forests between 0% and 10% the ice sheet starts to develop. With 10% forests the ice sheet only starts to grow with 500 ppmv CO_2, which is the lowest value for CO_2 in this test. It is very unlikely that there would be no ice below 500 ppmv CO_2.[1] We therefore conclude that the impact of albedo change due to forests is large, which is in accordance with previous work.[10] In order to test if it is warm enough for the 10% case to sustain forests, we have to look what the mean temperature of East Antarctica is of the warmest month just before glaciation, in most cases the Februari temperature. Figure 17 shows the result. The figure shows that the mean temperature in Februari of East Antarctica as calculated by the ZEBCM before glaciation is approximately 8 °C. This is around the threshold for forrests to be able to exist.[11] As a result, there may have been forests in small quantities. Those forests would be present at the margin of the continent, due to the lower altitude. When the ice sheet grows, the temperature drops very rapidly below the threshold of

8 °C, causing the forests to dissapear. We conclude however, that there may have been small percentages of the surface covered with forests in Antarctica before glaciation due to the high temperature of the warmest month.

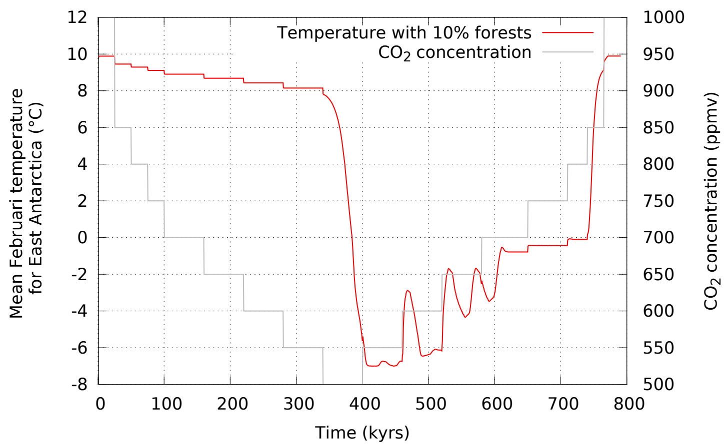

Figure 17: *Mean temperature of Februari in East Antarctica on the left axis and stepwise CO_2 concentration on the right axis. We assume that 10% of the continent is covered with forests.*

When the glaciation starts, the EAIS grows very rapidly. This is because of a feedback loop that takes place. We assume that all vegetation is gone at places where ice starts to grow, so the low albedo of forests is than suddenly gone, causing lower temperatures and intensifying the ice growth.

To test which components of the mass balance are important during ice growth, we show the accumulation, melt and total mass balance, combined with the modelled δ^{18}O. Figure 18 shows that the accumulation and melt are both large, but during times of ice growth (when the δ^{18}O increases), accumulation is larger than the melt. This results in a positive mass balance in the order of 1500 Gigaton/year.

In order to test when the ice sheet starts to grow with Antarctica completely covered with forests, we let the CO_2 concentration vary as before, but now between 600 ppmv and 200 ppmv. 200 ppmv CO_2 is close to the concentration of the last glacial maximum. So if the ice sheet does not grow under these circumstances, it will never grow under realistic circumstances with Antarctica completely covered with forests. Figure 19 shows that the EAIS does not develop with Antarctica covered with forests. Figure 20 shows that the mean temperature of Antarctica in Februari never gets below 8 °C. Februari is usually the warmest month in Antarctica, so it is warm enough to sustain forests at the margins of the continent. It is however, not realistic that it would be so warm with no ice with such low CO_2 concentrations.

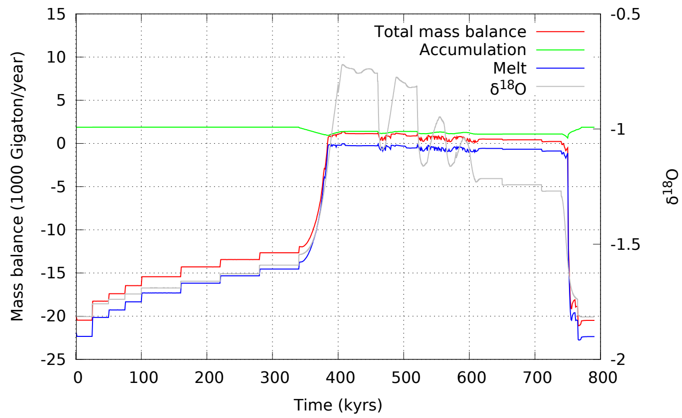

Figure 18: *Mass balance components of the East Antarctic ice sheet, combined with $\delta^{18}O$. It is assumed that 10% of the continent is covered with forests (when no ice is present).*

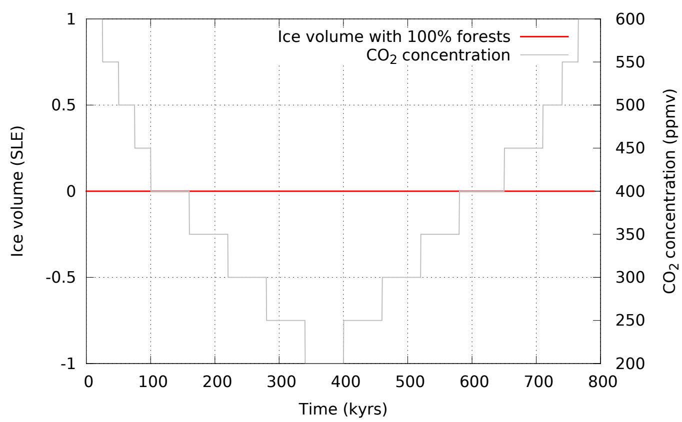

Figure 19: *East Antarctic ice volume on the left axis and stepwise CO_2 concentration on the right axis. The CO_2 concentration now varies between 600 and 200 ppmv. It is assumed that the entire continent is covered with forests.*

To test if it had been possible that Antarctica was for 20% covered with forests, we repeat the experiment with lower CO_2 concentrations. We observe in figure 21 that the EAIS starts to grow with 250 ppmv CO_2 if the Antarctic surface is covered with 20% forests. So if the surface is covered with 20% forests, the ice sheet still develops. However, It is very unlikely that the glaclial inception only takes place at such low CO_2 concentrations.[1,2] Therefore we can say that although ice still develops with 20% forests, it is not realistic. The determined CO_2 concentrations of glacial inception for 20% is far off the estimates of previous work.[1,2]

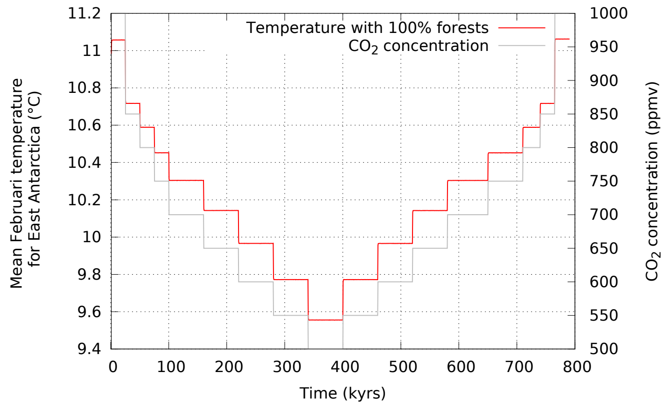

Figure 20: *Mean temperature of Februari in East Antarctica on the left axis and stepwise CO_2 concentration on the right axis. It is assumed that the entire continent is covered with forests.*

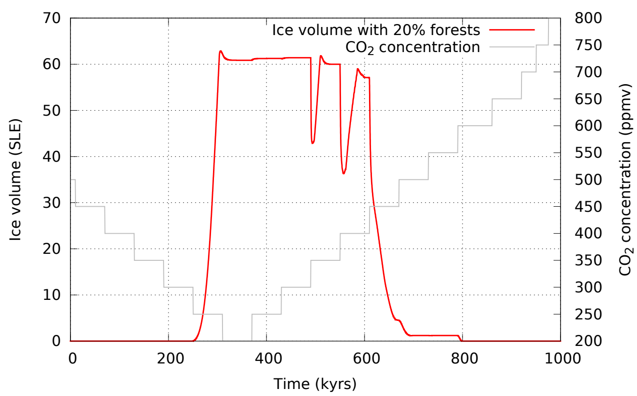

Figure 21: *East Antarctic ice volume on the left axis and stepwise CO_2 concentration on the right axis. It is assumed that 20% of the continent is covered with forests (when no ice is present).*

Figure 22 shows the temperature for each latitudinal band in the year of glacial inception. There is a large difference between the Februari temperature and the July temperature. The temperature difference between Februari and July at -90° is almost 50 °C. In Februari, the temperature starts to increase from -65° to -90°. This may be due to the increased insolation for higher latitudes during this month and the decreasing albedo during summer. Figure 22 shows that the temperature in the Antarctic region (-90° to -65°) in the summer is high enough in some places to sustain forests, although it is unlikely that the centre of the continent is so much warmer than the margins.

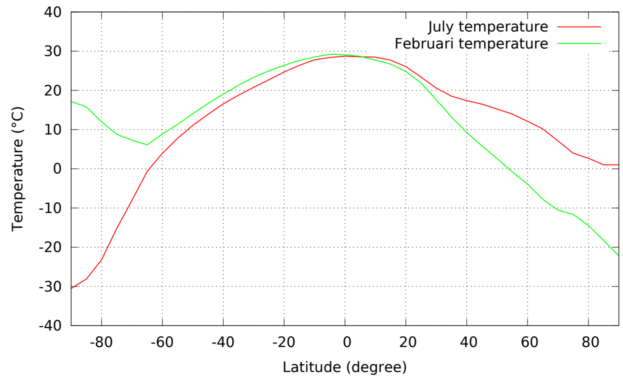

Figure 22: *Februari and July temperature per latitude just before glacial inception. This figure shows that the temperature difference between Februari and July of the Antarctic latitudinal bands is large.*

In summary, we conclude that vegetation has a large impact on glaciation. In order for the glacial inception to be in agreement with previous work,[1,2] the continent should not be covered with more than 10% forests. 20% Forest cover produces an ice sheet, but only for low CO_2 concentration. If there were forests, they would most likely be at the margins of the continent. Our results are in agreement with previous work concerning modelling of vegetation during the Eocene-Oligocene boundary.[10]

5 Discussion and conclusion

In order to reconstruct glacial inception of Antarctica during the Eocene-Oligocene transition, we have used a coupled one-dimensional Ice Sheet Model and a Zonally Averaged Energy Balance Climate Model.[8] We have shown that it models the glaciation cycles correctly. Variations in the CO_2 concentration cause large fluctuations in ice volume, therefore we have kept the CO_2 concentration constant with a stepwise function. We have also studied the mass balance sensitivity. We conclude that it has a large impact, and that the parameter C_{abl} = -30 produces the best results. We have also seen that fluctuations in ice volume of the EAIS are caused by the bedrock and not by sea level changes. The bedrock deforms when an ice sheet is on top of it. How fast the bedrock deforms is determined by the bedrock relaxation time τ_b, with τ_b = 3000 years producing the most plausible results.[1,2] Vegetation is an important factor in the climate system. Only small parts of Antarctica need to be covered with forests in order to substantially delay glaciation. In most cases, the temperature of the warmest month is high enough to sustain forests. However, it is less likely for Antarctica to have more than 10% of the surface covered with forests during times of glacial inception, because the CO_2 concentration would drop too low.

The kind of vegetation that was present and for which fraction of the land cover during the Eocene-Oligocene transition, is still uncertain. It is difficult to find evidence of Antarctic vegetation, due to the present ice sheet that covers the sediment layers. As a consequence, the best we can do is model the vegetation and see what influence it has on the glacial inception. If we compare this with sea records, it may be possible to say something about the vegetation that was present during first glacation. In one of our tests, we assumed that all of Antarctica was covered with forests, or covered with grass. This is of course not what happens in nature, but it should provide us some information about the importance of vegetation variability on the climate system. We also assumed that the Antarctic latitudinal bands (65° - 90°) all have the same ratio forests/grass. In reality we would expect that the amount of forests is lower at 90° and higher at 65°, which will facilitate ice inception. Our approximated ratio forests/grass is the mean over those latitudinal bands.

Many other experiments should be done to fully understand ice growth around the Eocene-Oligocene boundary. One of them is testing the ocean overturning. The overturning is already incorporated in the coupled model. Ocean overturning is known to change when the climate on Earth changes. The ocean cirulation was also different because of the continental configuration.[3] To better understand the extent of the overturning, it is possible to change the strength of the circulation. However, this has already been done for the last 800 kyrs,[8] and it proves to play only a minor role in the climate system in that time period.
Another experiment to do in order to improve the model, would be the atmospheric lapse rate. In our experiments it is kept constant, but in nature it is not. It can vary per location and it can vary in time. Further research is needed to determine how the lapse rate changes over time and per location.
The estimated values, calculated by the model for the bedrock deformation and relaxation time, can be compared with measurements of the present, in order to determine if they are physically correct.
The effect of clouds on the climate system is still uncertain. Clouds have a cooling, as well as a heating effect on the Earth. The albedo of clouds is usually high, reflecting much of the incoming short wave radiation back to space. Clouds also reflect long wave radiation back to Earth. They could therefore be a subject for experiments. Especially because it has been suggested that an increase in clouds increase the planetary albedo over highly reflective surfaces.[13] This is not taken account in this study.

In summary, we conclude that changes in the ablation parameter C_{abl} have a large impact on the mass balance. We have tested multiple values for this parameter with a stepwise CO_2 concentration, and concluded that C_{abl} = -30 produces the best result if we compare it with previous results.[1,2] We have shown that bedrock deformation has a large influence on the mass balance. If we do not take isostasy into account, the EAIS does not show large variations during the decay anymore, and it also takes longer for the ice sheet to dissapear, causing a larger hysteresis. As a consequence we conclude that bedrock deformations cause large fluctuations in ice volume, when the ice sheet first starts to melt. We also conclude that sea level changes are not as important as bedrock deformation. It is important to know how fast the bedrock deforms. Therefore, we have determined the bedrock relaxation time. We have estimated τ_b to be around 3000 years, which produces the most comparable result.[1,2] We have also seen that bigger values for τ_b cause large fluctuations and smaller values cause small fluctuations. The impact of vegetation on the climate system is large, especially that of forests. We have shown that only small percentages of the Antarctic continent needs to be covered with vegetation, in order to delay glaciation significantly, hence the effect of forests on the albedo is large. The mean temperature of East Antarctica of the warmest month before glaciation is around the threshold for forests to exist. Therefore forests may grow at the margin of the continent. If 10% of the continent is covered with forests, the CO_2 concentration of first glaciation is still in accordance with previous work.[1,2] We conclude that it may be possible that forests were present during first glaciation in small amounts, most likely at the margins of Antarctica. Nevertheless, vegetation has a very large impact on the climate system.

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